Section 16
Guidelines for Evaluating
Potentially Unstable Slopes and Landforms

PART 1. INTRODUCTION

This section of the board manual provides guidelines to evaluate potentially unstable slopes and landforms. It can be used to determine if additional information or a detailed environmental statement will be required before the submittal of a forest practices application for timber harvest or the construction of roads, landings, gravel pits, rock quarries, or spoil disposal areas on potentially unstable slopes or landforms that have the potential to deliver sediment or debris to a public resource or have the potential to threaten public safety.

It begins with an overview of the forest practices rules for potentially unstable slopes, which unstable landforms that are of concern, and the effects of landslides. Also included are important tools and concepts that can be used to determine if slopes are potentially unstable, descriptions of rule-identified unstable slopes and landforms, information and guidance on how to identify potentially unstable slope situations, the influence of forest practices activities on slope stability, and how to determine if delivery of material to public resources could occur. If you need to hire a qualified expert, guidelines for the contents of the expert report are listed at the end of this Section.

PART 2. OVERVIEW

Landslides occur naturally in forested basins and are an important process in the delivery of wood and gravel to streams. Wood and gravel play significant roles in creating stream diversity that is essential for fish use as habitat and spawning grounds. In the past, forest practices-caused landslides accelerated the processes that promote balance in nature creating a catastrophic regime that has contributed to the threatened and endangered status of certain species, as well as endangering human life in some instances. The forest practices rules are intended to protect public resources and public safety. The rules apply where there is *potential* for sediment and debris to be delivered to a stream, lake, other fish or wildlife habitat, domestic water supplies, or public capital improvements, or a threat to public safety. When the potential for instability is recognized, the likelihood that sediment and debris would travel far enough to threaten a public resource or public safety is considered. Many factors are part of that consideration including initial failure volume and nature of a landslide, landslide runout distance, and landscape geometry.

2.1 Potentially Unstable Landforms

Certain landforms are particularly susceptible to slope instability. Because of this, forest practices applications that propose activities on and near these landforms may be classified "IV-special" and receive additional environmental review under the State Environmental Policy Act (SEPA). Rule-identified unstable landforms that are described in this Section include bedrock hollows, convergent headwalls, and inner gorges with slopes >70% (35°), toes of deep-seated landslides with slopes >65% (32°), groundwater recharge areas for glacial deep-seated landslides, outer edges of meander bends, and other indications of slope instability. Below are descriptions of the types of landslide processes and consequences associated with them.

2.2 Landslide Types and Effects

Shallow landslides occur in bedrock hollows, convergent headwalls, and inner gorges with slopes >70%, on toes of deep-seated landslides with slopes >65%, and on the outer edges of meander bends. There are generally three types of shallow landslides: debris slides, debris flows, and hyper-concentrated floods. They are distinguished from each other by the ratio of water to solids contained in them.

Debris slides consist of aggregations of coarse soil, rock, and vegetation that lack significant water and move at speeds ranging from very slow to rapid down slope by sliding or rolling forward. The results are irregular hummocky deposits that are typically poorly sorted and non-stratified. Debris slides include those types of landslides also known as shallow rapid, soil slips, and debris avalanches. If debris slides entrain enough water they can become debris flows.

Debris flows are slurries composed of sediment, water, vegetation, and other debris. Solids typically constitute >60% of the volume (Pierson and Scott, 1985). Debris flows usually occur in steep channels, as landslide debris becomes charged with water (from soil water, or on entering a stream channel) and liquefies as it breaks up. These landslides can travel thousands of feet (or even miles) from the point of initiation, scouring the channel to bedrock in steeper channels.

Debris flows commonly slow where the channel makes a sharp bend and stop where the channel slope gradient becomes gentler than about 3°, or the valley bottom becomes wider and allows the flow to spread out.

Hyper-concentrated floods are flowing mixtures of water, sediment (dominantly sand-sized), and organic debris with solids that range between 20% and 60% by volume (Pierson and Scott, 1985). In forested mountains, they are commonly caused by the collapse of dams, such as those formed by landslide dams or debris jams (Figure 1). Impounded water and debris released when the dam is breached sends a flood wave down the channel that exceeds the magnitude of normal floods and generally extends beyond the range of influence that has been documented for debris flows (Johnson, 1991). Such hyper-concentrated floods can rise higher than normal rainfall- or snowmelt-induced flows along relatively confined valley bottoms, driving flood waters, sediment, and wood loads to elevations high above the active channel and, if present, the active floodplain.

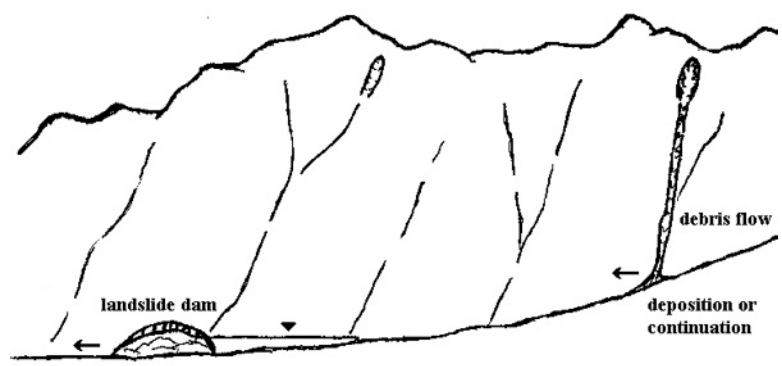

Figure 1. Debris flows, and hyper-concentrated floods

Debris flows and hyper-concentrated floods can occur in any unstable or potentially unstable terrain with susceptible valley geometry. In natural systems, debris flows and hyper-concentrated floods caused by dam-breaks are responsible for moving sediment and woody debris from hillslopes and small channels down into larger streams. But debris flows can also cause damage to streams by scouring channel reaches, disturbing riparian zones, impacting habitat and dumping debris onto salmonid spawning areas. Debris flows can cause elevated turbidity, adversely affect water quality downstream, threaten public safety, and damage roads and structures in their paths (Figure 2).

Figure 2. Road-initiated debris flows in inner gorges, Sygitowicz Creek, Whatcom County(Photo: DNR, 1983).

These debris flows coalesced and after exiting the confined channel at the base of the mountain, the new debris flow spread across a 1,000 foot wide swath for a distance of 2,000+ feet before entering the South Fork Nooksack River. Between the base of the mountain and the river the debris flow affected (if not severely damaged) a county road, farmyard and house sites, and 60+ acres of cultivated farm fields.

<u>2.3 Deep-Seated Landslides</u>

A more detailed explanation of deep-seated landslides is covered later in this section because deep-seated landslides are also landforms. Regardless of failure mechanism, deep-seated landslides are those in which the slide plane or zone of movement is well below the maximum rooting depth of forest trees (generally greater than three meters (10 feet)) and may extend to hundreds of feet in depth often including bedrock. Deep-seated landslides can occur almost anywhere on a hillslope and are typically associated with hydrologic responses in permeable geologic materials overlying less permeable materials. The larger deep-seated landslides can usually be identified from topographic maps or air-photos.

Certain key areas of deep-seated landslides may be sensitive to forest practices. The bodies and toes of deep-seated landslides are made up of incoherent collapsed material weakened from previous movement and therefore may be subject to debris slide and debris flow initiation in response to harvest or road building. Sediment delivery from shallow landslides on steep stream-adjacent toes of deep-seated landslides and steep side-slopes of marginal streams on the bodies of deep-seated landslides is common.

PART 3. MEASUREMENT OF SLOPE ANGLES

Slope gradients are commonly expressed in two different but related ways, as degrees of arc or percent rise to run. It is important to understand the relationships between them.

3.1 Degrees

A circle is divided into 360° of arc. Each degree is further divided into 60 minutes (60'), and each minute into 60 seconds (60"). The quadrant of the circle between a horizontal line and a vertical line comprises 90° of arc (Figure 3a).

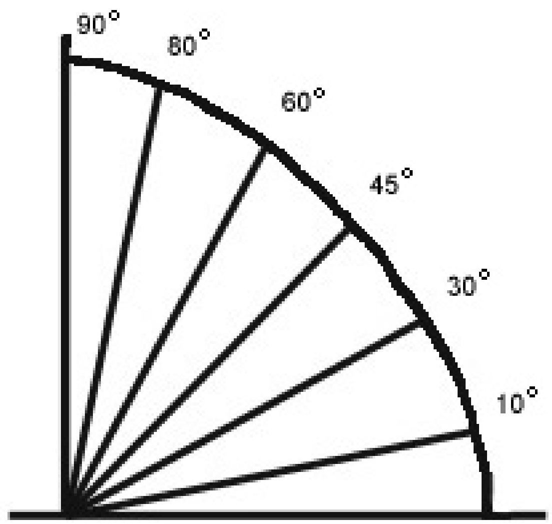

Figure 3a. Angles in degrees.

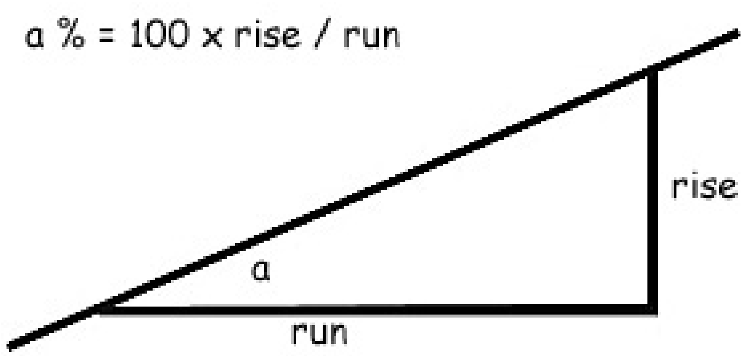

Figure 3b. Angles in percent.

3.2 Percent

In 3b, the horizontal distance between two points (distance between the points on a map) is called the run. The vertical distance (difference in elevation) is called the rise. The gradient can be expressed as the ratio of rise divided by run, a fraction that is the tangent of angle α . When multiplied by 100, this fraction is the percent slope.

3.3 Relationship of Degrees to Percent

Because of the differences in the ways they are calculated, each of these two slope measurements is better for certain applications. Because it is more precise at gentle slopes, percent is best for measuring and expressing small angles, such as the gradients of larger streams. But for steeper slopes, the constant angular difference and smaller numbers (an 85° slope is 1143%) make degrees more useful.

Figure 4 shows approximate equivalences for gradients expressed in degrees and percent. Note that there is a rough 2:1 ratio in the 30 to 40° range (e.g., 35° = 70% slope), but beware - this relationship changes dramatically at gentler and steeper angles.

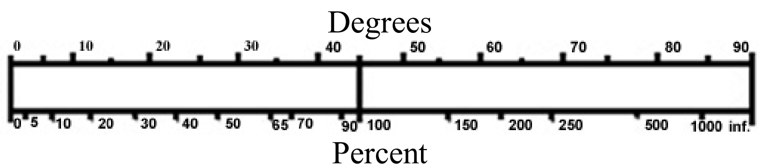

Figure 4. Slope gradients in degrees and percent.

PART 4. SLOPE FORM

Slope shape is an important concept when considering the mechanisms behind shallow landsliding. Understanding and recognizing the differences in slope form is key in potentially unstable landform recognition. There are three major slope forms to be observed when looking across the slope (contour direction): divergent (ridgetop), planar (straight), and convergent (spoon-shaped) (Figure 5a). Landslides can occur on any of these slope forms but divergent slopes tend to be more stable than convergent slopes because water and debris spread out on a divergent slope whereas water and debris concentrate on convergent slopes. Convergent slopes tend to lead into the stream network, encouraging delivery of landslide debris to the stream system. Planar slopes are generally less stable than divergent slopes but more stable than convergent slopes. In the vertical direction, ridgetops are convex areas (bulging outward) and tend to be more stable than planar (straight) mid-slopes and concave areas (sloping inward) (Figure 5b).

Additionally, slope steepness can play a significant role in shallow landsliding. Steeper slopes tend to be less stable. The soil mantle, depending upon its make-up, has a natural angle at which it is relatively stable (natural angle of repose). When hillslopes evolve to be steeper than the natural angle of repose of the soil mantle, the hillslope is less stable and more prone to shallow landslides, especially with the addition of water. The combination of steep slopes and convergent topography has the highest potential for shallow landsliding.

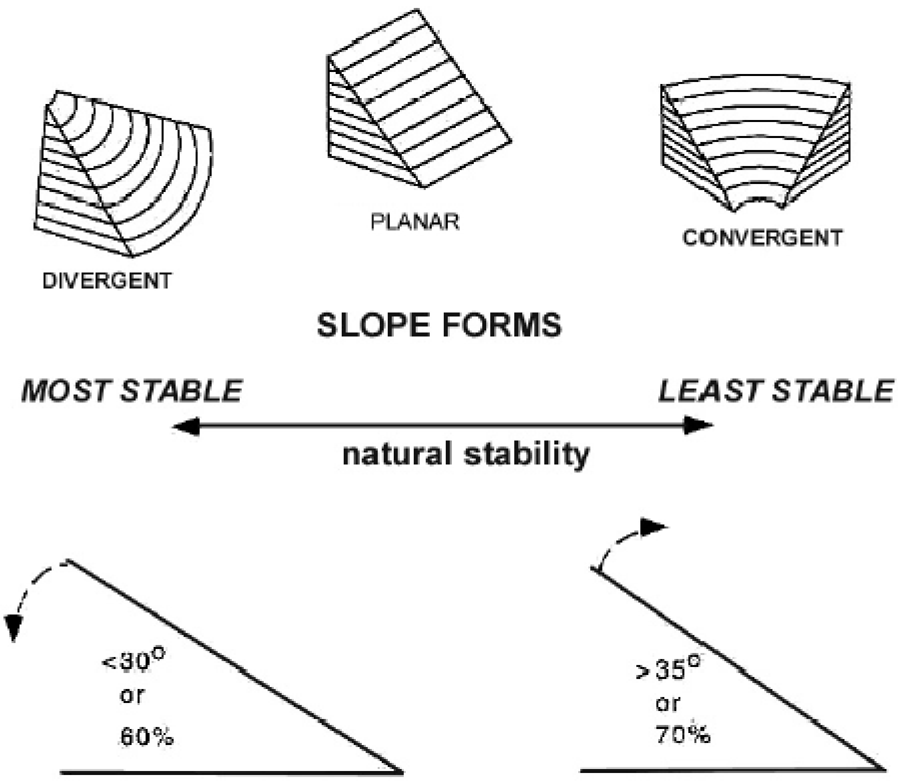

Figure 5a. Slope configurations as observed in map view.

This figure shows three major slope forms (divergent, planar, and convergent) and their relative stability. These slope form terms are used in reference to contour (across) directions on a slope. Convergent areas with slope greater than 35° (70%) are the most shallow landslide-prone (Benda, et al, 1997/8).

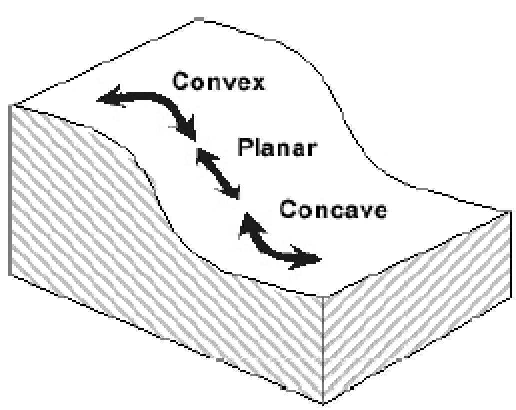

Figure 5b. Slope configurations as observed in profile: convex, planar, and concave. These terms are used in reference to up and down directions on a slope (Drawing: Jack Powell, DNR, 2004).

PART 5. DESCRIPTION OF UNSTABLE AND POTENTIALLY UNSTABLE LANDFORMS AND PROCESSES

Areas of unstable landforms can usually be identified with a combination of topographic and geologic maps, aerial photographs, watershed analysis mass wasting map unit (MWMU) maps, landslide-hazard maps from the Regional (Unstable) Landform Identification Project (RLIP), Landslide Hazard Zonation Project (LHZ), and modeled slope stability morphology (SLPSTAB, SHALSTAB, SINMAP) output maps. However, field observation is normally required to precisely delineate landform boundaries, gradients, and other characteristics. In most instances, landform terms described herein are also used in the scientific literature. For the purposes of Washington forest practices, the rule-identified landform terms, definitions, and descriptions supercede those used in the scientific literature. Note that all sizes, widths, lengths, and depths are approximate in the following discussion of unstable landforms, and are not part of the rule-identified definitions. Sizes are included to help visualize the landforms.

5.1 Bedrock Hollows, Convergent Headwalls, Inner Gorges

These three landforms are commonly associated with each other as shown in Figures 6 and 7.

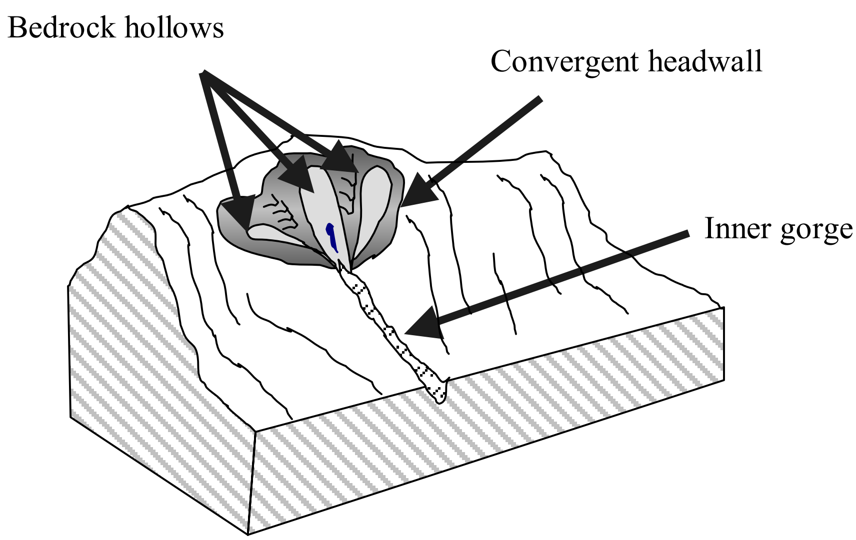

Figure 6. Typical hillslope relationships between bedrock hollows, convergent headwall, and inner gorge (Drawing: Jack Powell, DNR, 2003).

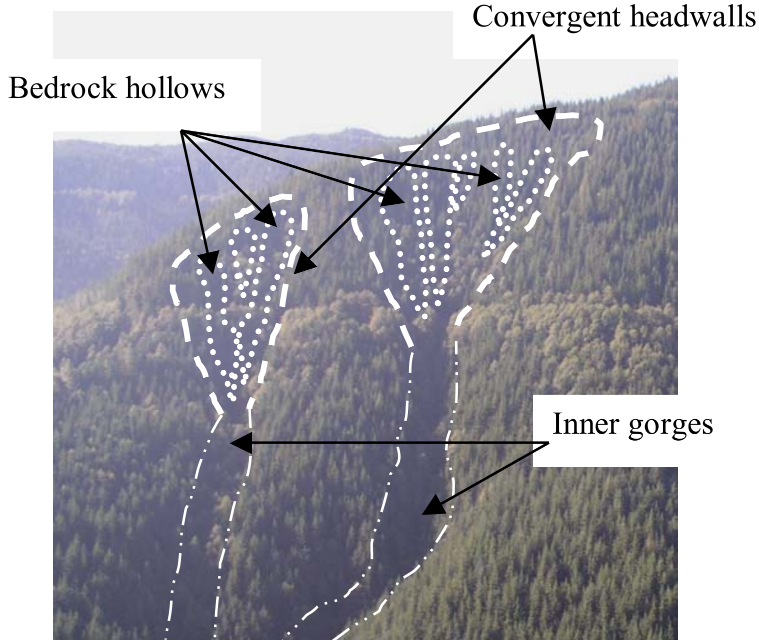

Figure 7. Common hillslope relationship: bedrock hollows in convergent headwalls draining to inner gorges (Photo and drawing: Jack Powell, DNR, 2003).

Bedrock hollows are also called colluvium-filled bedrock hollows, zero-order basins, swales, bedrock depressions, or simply hollows. Not all hollows contain bedrock so the term “bedrock” hollow can be a misnomer. However, the forest practices rules cite these features as “bedrock” hollows so this is the term used in the Board Manual. Hollows are commonly spoon-shaped areas of convergent topography with concave profiles on hillslopes. They tend to be oriented linear up- and down-slope. Their upper ends can extend to the ridge or begin as much as several hundred feet below ridge line. Most hollows are approximately 75 to 200 feet wide at their apex (but they can also be as narrow as several feet across at the top), and narrow to 30 to 60 feet downhill. Hollows should not be confused with other hillslope depressions such as small valleys, sag areas (closed depressions) on the bodies of large deep-seated landslides, tree wind-throw holes (pit and mound topography), or low-gradient swales.

Hollows often form on other landforms such as head scarps and toes of deep-seated landslides. Bedrock hollows can occur singly or in clusters that define a convergent headwall. They commonly drain into inner gorges (Figure 8).

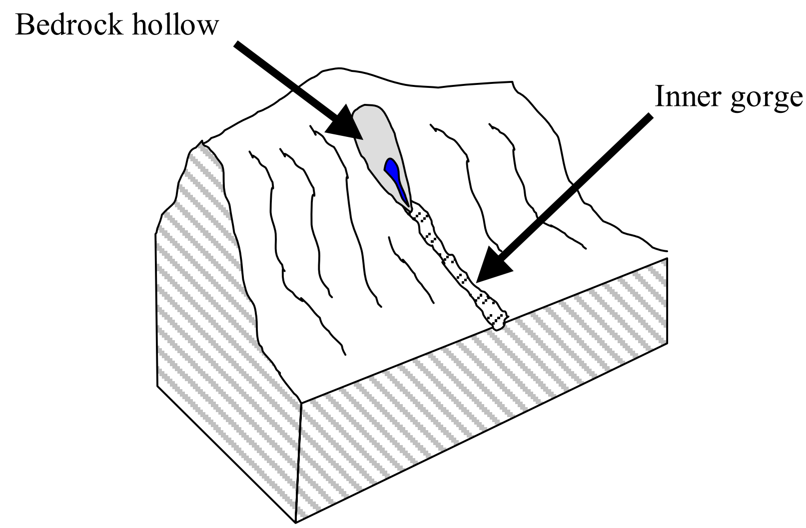

Figure 8. Bedrock hollow and relationship to inner gorge (Drawing: Jack Powell, DNR, 2003).

Hollows usually terminate where distinct channels begin. This is at the point of channel initiation where water emerges from a slope and has carved an actual incision. Steep bedrock hollows typically undergo episodic evacuation of debris by shallow-rapid mass movement, followed by slow refilling with colluvium that takes years or decades. Unless they have recently experienced evacuation by a landslide, hollows are partially or completely filled with colluvial soils that are typically deeper than those on the adjacent spurs and planar slopes. Recently evacuated hollows may have water flowing along their axes whereas partially evacuated hollows will have springs until they fill with sufficient colluvium to allow water to flow subsurface.

Figure 9 illustrates the evolution of a bedrock hollow. Drawing “a” shows that over a period of tens to hundreds or thousands of years in some places, sediment accumulates in a hollow. When the soil approaches a depth of 3 to 5 feet (1-2 meters), the likelihood of landslides increases. Recurrent landsliding within the hollow slowly erodes bedrock and maintains the form of the hollow (Drawing “b”). After a landslide, bedrock is exposed (and also seeps or springs) and the risk of additional sliding is reduced, but not gone. Drawing “c” shows soil from the surrounding hillsides (colluvium) slowly re-filling the hollow. Roots help stabilize the soil.

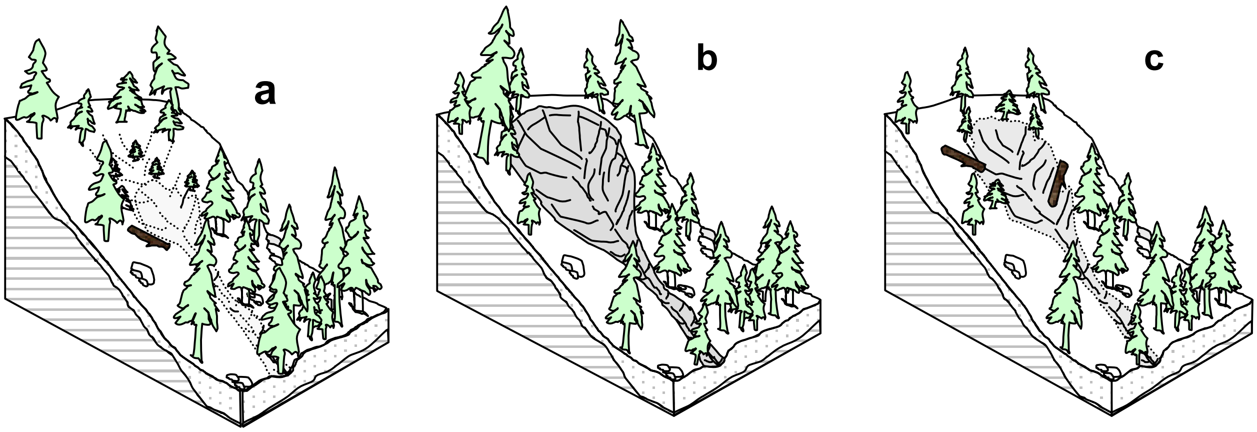

Figure 9a-c. Evolution of a bedrock hollow following a landslide (adapted from Dietrich et al., 1988 by Jack Powell, DNR, 2004).

The common angle of repose for dry, cohesionless materials is about 36° (72%), and saturated soils can become unstable at lower gradients. Thus, slopes steeper than about 35° (70%) are considered susceptible to shallow debris slides. "Bedrock" hollows are formed on slopes of varying steepness. Hollows with slopes steeper than 70% (approximately 35°) are potentially unstable in well-consolidated materials, but hollows in poorly consolidated materials may be unstable at lower angles. *Note: Bedrock hollow slopes are measured on the steepest part of the slope generally not along the axis unless the hollow is full (Figure 10).*

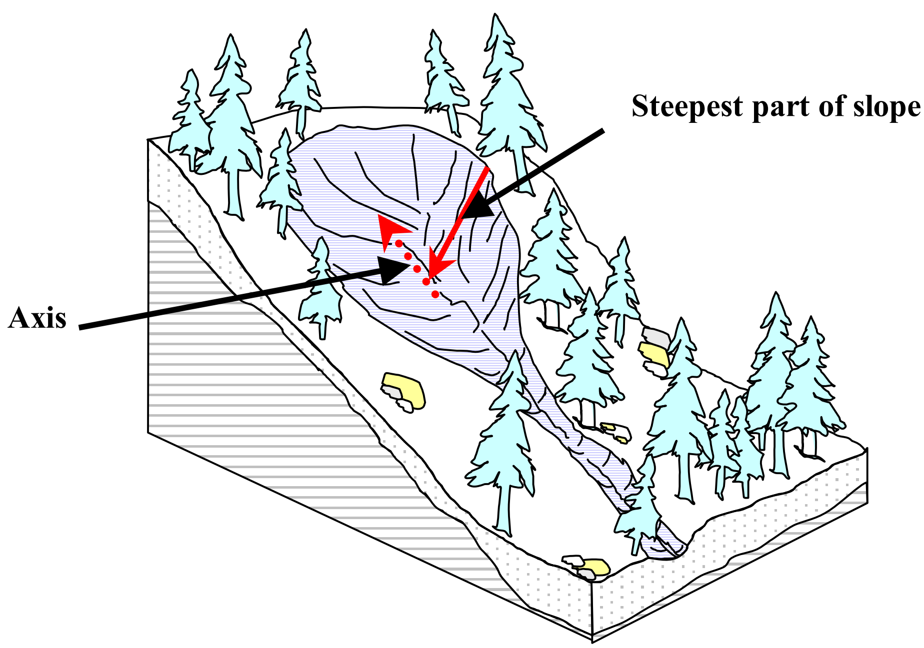

Figure 10. Bedrock hollow slopes are measured on the steepest part of the slope generally not along the axis (Drawing: Jack Powell, DNR, 2004).

Vegetation can provide the critical cohesion on marginally stable slopes and removes water from the soil through evapotranspiration. Leaving trees in steep, landslide-prone bedrock hollows helps maintain rooting strength and should reduce the likelihood of landsliding (Figure 11). However, wind-throw of the residual trees following harvest can be associated with debris slide or debris flow events. In high wind environments, it is essential to harvest in a manner that will limit the susceptibility of the residual trees to wind-throw as well as to reduce the potential for landslides (for example leaving wider strips, pruning or topping trees in the strips, or feathering the edges of reserve strips).

Figure 11. Example of leave areas protecting unstable slopes (Photo: Venice Goetz, DNR, 2004).

Convergent headwalls are funnel-shaped landforms, broad at the ridgetop and terminating where headwaters converge into a single channel. A series of converging bedrock hollows may form the upper part of a convergent headwall (Figure 12). Convergent headwalls are broadly concave both longitudinally and across the slope, but may contain sharp ridges that separate the bedrock hollows or headwater channels (Figure 13a, b, and Figure 14).

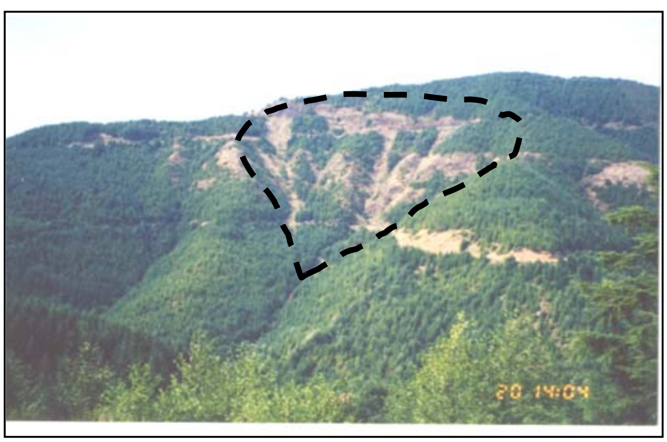

Figure 12. Convergent headwall example (Photo: Venice Goetz, DNR, 1995).

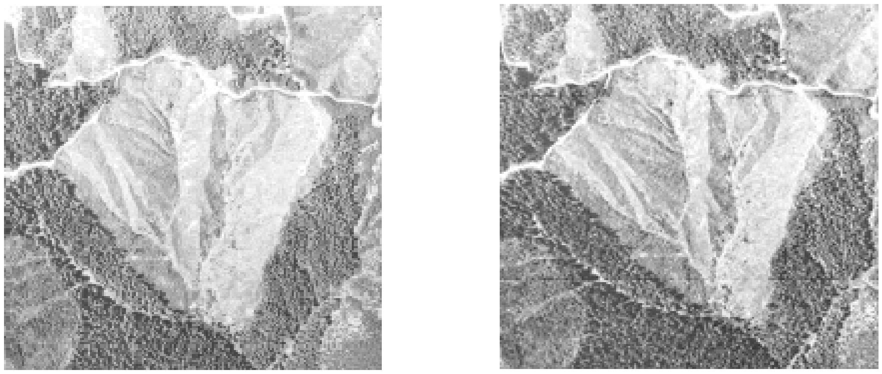

Figure 13a, b. Stereo-pair of a clearcut convergent headwall in Pistol Creek basin, North Fork Calawah River, Washington.

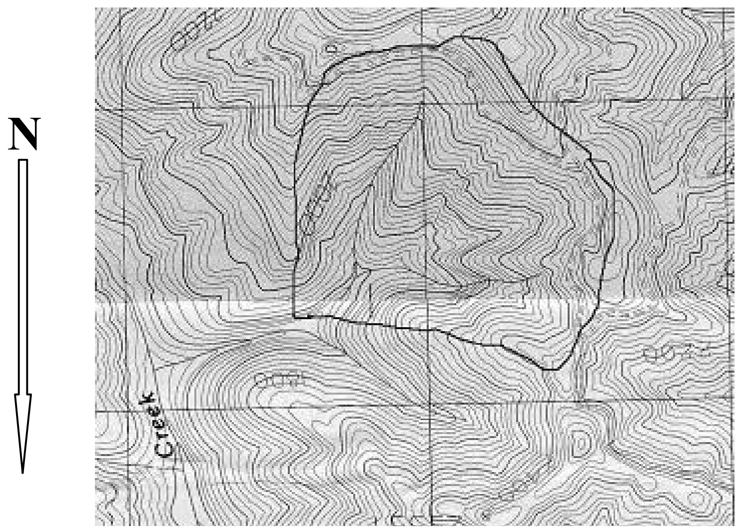

Figure 14. Topographic map and outline of convergent headwall displayed in the stereo-pair of Figure 13a, b. Scanned from portions of Hunger Mountain and Snider Peak USGS 7.5' quadrangles.

Convergent headwalls generally range from about 30 to 300 acres. Slope gradients are typically steeper than 35° (70%) and may exceed 45° (94%). Unlike bedrock hollows, which exhibit a wide range of gradients, only very steep convergent landforms with an obvious history of landslides are called convergent headwalls. Soils are thin because landslides are frequent in these landforms. It is the arrangement of bedrock hollows and first-order channels on the landscape that causes a convergent headwall to be a unique mass-wasting feature. The highly convergent shape of the slopes, coupled with thin soils (due to frequent landslides), allows rapid onset of subsurface storm water flow. The mass-wasting response of these landforms to storms, disturbances such as fire, and to forest practices activities is much greater than is observed on other steep hillslopes in the same geologic settings. Convergent headwalls may be also prone to surface erosion from the scars of frequent landslides.

Channel gradients are extremely steep within convergent headwalls, and generally remain so for long distances downstream. Landslides that evolve into debris flows in convergent headwalls typically deliver debris to larger channels below. Channels that exit the bottoms of headwalls have been formed by repeated debris flows and are efficient at conducting them. Convergent headwalls commonly have debris fans at the base of their slopes.

Inner gorges are canyons created by a combination of stream down-cutting and mass movement on slope walls. Inner gorges are characterized by steep, straight or concave side-slope walls that commonly have a distinctive break in slope (Figure 15). Debris flows, in part, shape inner gorges by scouring the stream, undercutting side slopes, and/or depositing material within or adjacent to the channel (Figure 16). Inner gorge side-slopes may show evidence of recent landslides, such as obvious landslides, raw un-vegetated slopes, young, even-aged disturbance vegetation, or areas that are convergent in contour and concave in profile. Because of steep slopes and proximity to water, landslide activity in inner gorges is highly likely to deliver sediment to streams or structures downhill. Exceptions can occur where benches of sufficient size to stop moving material exist along the gorge walls, but these are uncommon.

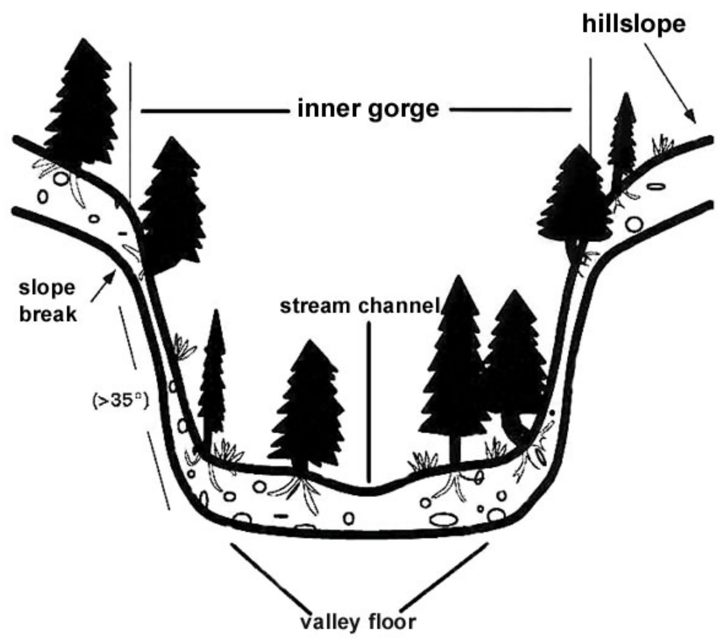

Figure 15. Cross-section of an inner gorge. This view emphasizes the abrupt steepening below the break-in-slope (Drawing: Benda, et al, 1998).

Figure 16. Photograph showing how debris flows help shape features related to inner gorges. (For example, over-steepened canyon wall, U-shaped profile, buried wood, distinctive break in slope along margins of inner gorge (Photo: Laura Vaugeois, DNR, 2004).

The geometry of inner gorges varies. Steep inner gorge walls can be continuous for great lengths, as along a highly confined stream that is actively down cutting, but there may also be gentler slopes between steeper ones along valley walls. Inner gorges can be asymmetrical with one side being steeper than the other. Stream-eroded valley sides, which can be V-shaped with distinct slope breaks at the top, commonly do not show evidence of recent landsliding as do inner gorges which tend to be U-shaped. In practice, a minimum vertical height of 10 feet is usually applied to distinguish between inner gorges and slightly incised streams.

The upper boundary of an inner gorge is assumed to be a line along the first break in slope of at least 10 º or the line above which gradients are mostly gentler than 35° (70%) and convex. The delineating break-in-slope occurs where over-steepened slopes related to inner gorge erosion processes intersect slopes formed from normal hillslope erosion processes. While the upper inner gorge boundary is typically distinct, in some places it can be subtle and challenging to discern. Inner gorge slopes tend to be especially unstable at the point where the slope breaks because the abrupt change in gradient causes subsurface water to collect within the soil matrix which can destabilize the soil mass and initiate movement. Just as for all other landforms, inner gorge slopes should be measured along the steepest portion of the slope (see Figure 10).

The steepness of inner gorges is dependent on the underlying materials. In competent bedrock, gradients of 35° (70%) or steeper can be maintained, but soil mantles are sensitive to root-strength loss at these angles. Slope gradients as gentle as about 28° (53%) can be unstable in gorges cut into incompetent bedrock, weathered materials, or unconsolidated deposits.

Erosion along the gorge walls can intercept shallow groundwater forming seeps along the sides of the inner gorge, which promotes continued mass wasting. Root strength along walls and margins of inner gorges has been found to be a factor that limits the rates of mass wasting. Inner gorge areas can lose root strength when trees blow down. However, downed timber has a buttressing effect providing some slope reinforcement. Effective rooting width of forest trees is approximately the same as the crown width. In some instances where the inner gorge feature is highly unstable it is necessary to maintain trees beyond the slope break. Use the rooting strength of trees adjacent to the landform for additional support.

5.2 Other Indicators of Slope Instability or Active Movement

In addition to the landforms above, other indicators of slope instability or active movement may include:

(a) topographic and hydrologic
- bare or raw, exposed, un-vegetated soil on the faces of steep slopes
- boulder piles
- hummocky or benched surfaces, especially below crescent-shaped headwalls
- fresh deposits of rock, soil, or other debris at the base of a slope
- ponding of water in irregular depressions or undrained swampy areas on the hillslope above the valley floor
- cracks in the surface (across or along slopes, or in roads)
- seepage lines or springs and soil piping
- deflected or displaced streams (streams that have moved laterally to accommodate landslide deposits)

(b) vegetational
- jack-strawed, back-rotated, or leaning trees
- bowed, kinked, or pistol-butted trees
- split trees
- water-loving vegetation (horsetail, skunk cabbage, etc.) on slopes
- other patterns of disturbed vegetation

No one of these indicators necessarily proves that slope movement is happening or imminent, but a combination of several indicators could indicate a potentially unstable site.

Additional information about landslide processes, unstable landforms, and the effects of forest practices on unstable landforms is available in *"A Guide for Management of Landslide-Prone Terrain in the Pacific Northwest"* by the British Columbia Ministry of Forests (1994); and Hillslope Stability and Land Use (Sidle et al, 1985).

Deep-seated landslides are those in which the slide plane or zone of movement is well below the maximum rooting depth of forest trees (generally greater than 10 feet or 3 meters). Deep-seated landslides may extend to hundreds of feet in depth, often including bedrock. Deep-seated landslides can occur almost anywhere on a hillslope and can be as large as several miles across or as small as a fraction of an acre. The larger ones can usually be identified from topographic maps or aerial photographs. Many deep-seated landslides occur in the lower portions of hillslopes and extend directly into stream channels whereas deep-seated landslides confined to upper slopes may not have the ability to deposit material directly into channels.

One common triggering mechanism of deep-seated landslides results from the over-steepening of the toe by natural means such as glacial erosion or fluvial undercutting, fault uplift, or by human-caused excavations. Initiation of such landslides has also been associated with changes in land use, increases in groundwater levels, and the degradation of material strength through natural processes. Movement can be complex, ranging from slow to rapid, and may include small to large displacements.

Deep-seated landslides characteristically occur in weak materials such as thinly layered rocks, unconsolidated sediments, deeply weathered bedrock, or rocks with closely spaced fractures. Examples include: clay-rich rocks, such as the Lincoln Creek Formation of west-central Washington; thinly layered rocks, such as phyllite in northwest Washington; and deeply weathered volcanic rocks that cover the Willapa Hills of southwest Washington. Deep-seated landslides can also occur where a weak layer or prominent discontinuity is present in otherwise strong rocks, such as clay or sand-rich interbeds in the basalts of eastern Washington or a fault plane or intersecting joint set. In northwest Washington and on the Olympic Peninsula, deep-seated landslides commonly occur along silt or clay beds that are overlain by sandy units such as glacial deposits.

There are three main parts of a deep-seated landslide: the scarps (head and side), along which marginal streams can develop; the body, which is the displaced slide material; and the toe, which also consists of displaced materials. The downslope edge of the toe can become oversteepened from stream erosion or from the rotation of the slide mass. A deep-seated landslide may have several of each of these parts because small deep-seated landslides can be found nested within larger slides. These three main parts are shown in Figures 17 and 18. The head- and side-scarps together form an arcuate or horseshoe shaped feature that represents the surface expression of the rupture plane. The body and toe area are usually hummocky and the flow path of streams on these landslide sections may be displaced in odd ways due to differential movement of landslide blocks. The parts of deep-seated landslides that are susceptible to shallow landslides and potential sediment delivery are steep scarps (including marginal stream side slopes) and toe edges.

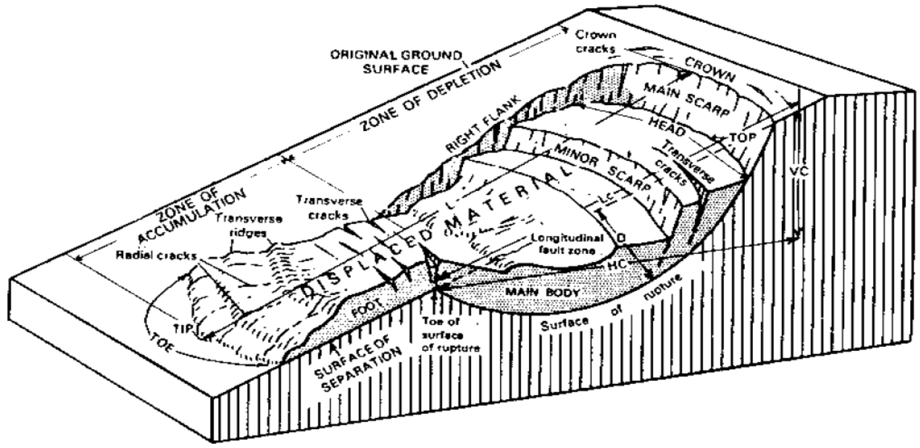

Figure 17. Rotational deep-seated landslide. Rotational displacement of blocks of soil commonly occur at the head of the landslide. Slow flow (an earthflow) may be found at the toe (Drawing: Varnes, 1978).

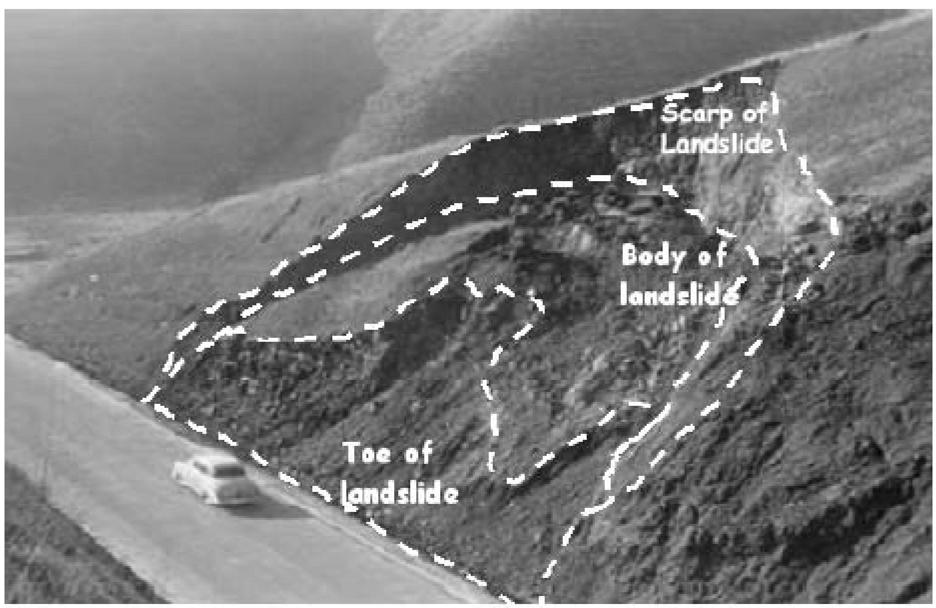

Figure 18. Deep-seated landslide showing the head scarp, side-scarps, body, and toe. Some of the toe has been removed in building and maintaining the highway (adapted from USGS photo).

The sensitivity of any particular landslide to forest practices is highly variable. Deep-seated scarps and toes may be over-steepened and streams draining the displacement material may be subject to debris slide and debris flow initiation in response to harvest or road building. Movement in landslides is usually triggered by accumulations of water at the slide zone, so land-use changes that alter the amount or timing of water delivered to a landslide can start or accelerate movement. Generally, avoiding the following practices will prevent most problems: destabilizing the toe by the removal of material during road construction or quarrying; overloading the slopes by dumping spoils on the upper or mid-scarp areas, or compacting the soil in these places which could change subsurface hydrology; and directing additional water into the slide from road drainage or drainage capture. The loss of tree canopy interception of moisture and the reduction in evapotranspiration through timber removal may also initiate movement of the slide.

5.3 Toes of Deep-Seated Landslides

The toes of deep-seated landslides are a forest practices regulatory landform. In this context "deep-seated landslide toes" means the down slope toe edges, not the entire toe area of displacement material (see Figure 17). Landslides that have toe edges adjacent to streams have a high potential for delivery of sediment and wood to streams. In such situations, streams can undercut the landslide toes and promote movement. Such over-steepened toes of deep-seated landslides can also be sensitive to changes caused by harvest and road construction. Resulting instability can take the form of shallow landslides, small-scale slumping, or reactivation of parts or the whole of deep-seated landslide. Because deep-seated landslides are usually in weak materials (further weakened by previous movement), an angle of 33° (65%) is the threshold value used on the potentially unstable toe edges. Regardless of the surface expression of the toe, it is best to avoid disrupting the balance of the landslide mass by cutting into or removing material from the toe area.

5.4 Groundwater Recharge Areas of (Glacial) Deep-Seated Landslides

Groundwater recharge areas of deep-seated slides are located in the lands up-slope that can contribute subsurface water to the landslide. In some cases this can include upslope portions of the landslide itself. Cemented soil horizons, fine-grained soils, and/or the presence of glacial till can be factors controlling the infiltration and flow of groundwater (Bauer and Mastin, 1997; Vaccaro et al., 1998). Groundwater perching and the characteristics of the overlying groundwater recharge area can be important factors in a deep-seated failure, especially for landslides in glacial sand and other unconsolidated sequences that overlie glacial-lake clay deposits or till (Figure 19). This is a common configuration of the glacial deposits in much of the northern half of western Washington (e.g., landslides in Seattle (Gerstel and others, 1997) and in the Stillaguamish River valley (Benda and others, 1988)), but this type of landslide also occurs in alpine glacial deposits in southwest Washington far from the mountain front. Groundwater filtering down through porous sand layers is trapped above the poorly permeable clay or till. During storm events, the sand above the clay becomes saturated creating a buoyant effect and lowering cohesion in the sand, both of which weaken the contact between the clay and sand. This in turn causes the overlying mass to slide along the sand/clay contact. A key predictive observation is noting the presence of a horizontal line of springs (groundwater refluxing) or a line of vegetation at the contact between the permeable and less permeable layers. Land uses such as poorly planned ditches or large-scale, even-aged harvesting that alter the timing or volumes of groundwater recharge in the slide zone can start or accelerate landslide movement.

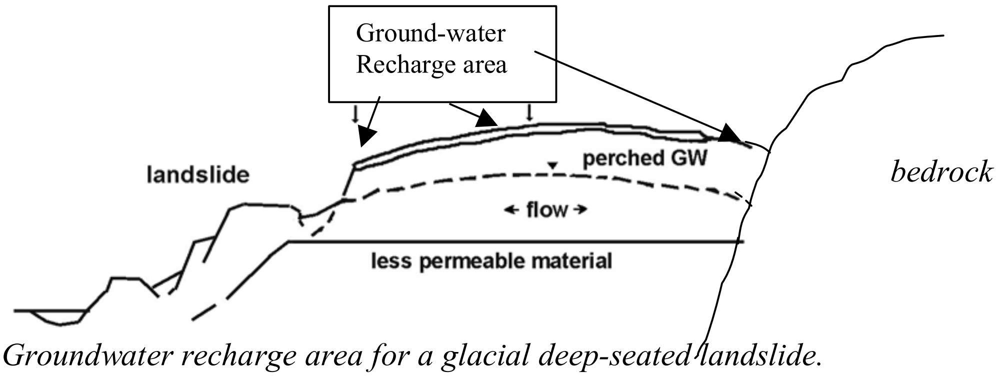

Figure 19. Groundwater recharge area for a glacial deep-seated landslide.

Because of the likelihood of subsurface water flow along perching layers in glacial strata, recharge areas for glacial deep-seated landslides may be classified “IV-special” under the forest practices rules and require further investigation. Therefore, it is important to characterize groundwater recharge areas and local stratigraphy in terms of an evaluation of the potential for changes in the water balance due to forest practices activities and an assessment of the degree to which a potential hydrologic change can be effectively delivered to a glacial deep-seated landslide. In the absence of other information, the recharge area is assumed to be equivalent to the surface (topographically defined) basin directly above the active slide. A more refined estimate of the spatial extent of a groundwater recharge area can be interpreted from field observation of soil profiles, stratigraphy, logs of wells or boreholes, or large-scale geologic maps.

5.5 Outer Edges of Meander Bends

Streams can create unstable slopes by undercutting the outer edges of meander bends along valley walls or high terraces of an unconfined meandering stream (Figure 20). The outer edges of meander bends are susceptible to shallow landsliding including debris avalanching and small-scale slumping, and deep-seated landsliding. The outer edges of meander bends may be protected by the riparian management zone (RMZ) or channel migration zone (CMZ) rules if the slopes are not particularly high and are contained within the riparian leave areas or within the CMZ (see Board Manual Section 2). As with other situations of overlapping forest practices rules, the harvest unit layout should reflect the extent of the greater of the protections.

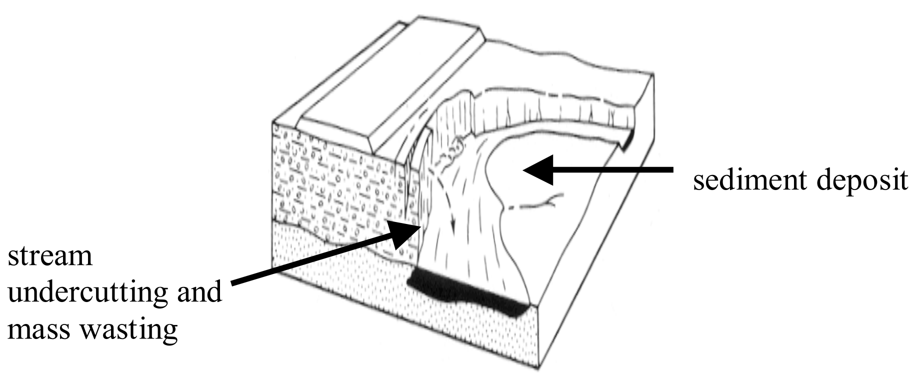

Figure 20. Outer edge of a meander bend showing mass wasting on the outside of the bend and deposition on the inside (adapted from Varnes, 1978).

PART 6. POTENTIALLY UNSTABLE LANDFORMS AND MATERIALS-REGIONAL LISTS

The Regional (Unstable) Landform Identification Project (RLIP) is a result of the Forests and Fish Report and is being conducted statewide at this time. The purpose of the RLIP is to note and validate region-specific unstable landforms that are not included in the present forest practices rules so that known unstable landforms are not overlooked during the forest practices application process. The final products will be in the form of short reports (validations) and maps that describe (generally and specifically) and locate these regional unstable landforms. This information is intended to be used as a screening tool for forest practices applications and may be eventually included in the forest practices rules and this Board Manual Section.

PART 7. DELIVERY

Landslides occur naturally in forested basins and are an important process in the delivery of wood and gravel to streams. Wood and gravel play important roles in creating stream diversity that is essential for fish use as habitat and spawning grounds. In the past, landslides as a result of forest practices activities have created a catastrophic regime that has contributed to the threatened and endangered status of certain species, as well as endangering human life in some instances. The rules apply where there is *potential* for sediment and debris to be delivered to a public resource or threatens public safety. When the potential for instability is recognized, the likelihood that sediment and debris would travel far enough to threaten a public resource or public safety should be considered. Many factors are part of that consideration including the initial failure volume of a landslide, the runout distance of a landslide, and landscape geometry.

It is difficult to prescribe guidelines for delivery distances because each situation has a special combination of process and topography. Deep-seated landslides can move anywhere from a few inches to a few miles depending on the friction of the slip plane, the forces pulling the landslides down, and the shear strength resisting those forces. Larger landslides are more likely to be able to move great distances at gentle gradients, but they are also less likely to be significantly affected by forest practices activities.

Timber harvest and road building can cause shallow landslides on steep slopes. Travel distances for such landslides depend on the amount of water contained in or entrained by them. Considering that rain, snowmelt, or some other extreme water inputs trigger the vast majority of landslides in the Pacific Northwest, it should be noted that almost all landslides contain some amount of water that tends to mobilize the soil or rock. Debris slides that do not reach streams (i.e., do not absorb large volumes of additional water) usually deposit their debris on the hillslope; and are typically unable to move far across large areas of flat ground. However, since most landslides occur during storm conditions, a large proportion of debris slides do reach flowing channels and create the opportunity to entrain enough water to become debris flows. These flows are quite mobile, and can travel great distances in steep or moderate gradient channels.

When channel gradients drop below 12° (20%), debris flows no longer scour and generally begin to slow down. On slopes gentler than about 3-4° (5-7%) debris flows commonly start losing their momentum and the solids entrained in them (rock, soil, organic material) tend to settle out.

Travel distance of a debris flow once it reaches a low-gradient surface is a function of its volume and viscosity. The solid volume of a debris slide or flow deposit is a function of soil depth, distance traveled down the hillslope, and the gradient of the traveled path. The proportion of water is the main control on viscosity. Field or empirical evidence should be used for determining the runout distance.

Even if the main mass of a landslide or debris flow comes to rest without reaching a public resource, there is the possibility that secondary effects may occur. Bare ground exposed by mass movement and disturbed piles of landslide debris can be chronic sources of fine sediment to streams until stabilized by revegetation. If flowing water (seepage, overland flow, or small streams) can entrain significant volumes of fine sediment from such surfaces, the possibility of secondary delivery must be evaluated, along with the likelihood of impact by the initial movement event itself.

To assess the potential for delivery and estimate runout distance, analysts can evaluate the history of landslide runout in the region, use field observations, and/or use geometric relationships appropriate from the scientific literature. In any situation where the potential for delivery is questionable, it is best to have a geotechnical expert examine the situation and evaluate the likelihood of delivery. If forest practices are to be conducted on an unstable landform with questionable or obvious potential to impact a public resource, a geotechnical report written by a qualified expert is required.

PART 8. GEOTECHNICAL REPORTS

When harvesting or building roads on potentially unstable slopes a geotechnical report is required to explain how the proposed forest practice is likely to affect slope stability, delivery to public resources, and public safety. The applicant must also submit to DNR a SEPA checklist and additional information as described in WAC 222-10-030.

Geotechnical reports must be prepared by **qualified experts** and must meet the requirements as described in WAC 222-10-030(5)

Effective July 1, 2002, qualified experts must be licensed with Washington's Geologist Licensing Board. For more information on the geologist licensing process, refer to WAC 308-15-010 through 308-15-150, or visit the Geology Board's web site at www.wa.gov/dol/bpd/geofront.htm. The education and field experience on forestlands will still be required, in addition to the appropriate geologist license.

8.1 Guidelines for Geo-Technical Reports

The following elements (a-f) should be included in geotechnical reports submitted by qualified experts:

(a) *Prepare an introductory section.* This section should describe the qualifications of the expert to ensure he/she meets the aforementioned requirements. It should also reference the forest practices application number (if previously submitted), the landowner(s) and operator(s) names, and a brief description of field trip(s) to the area, including dates, relevant weather conditions, and the locations visited.

(b) *Describe the geographic, geologic, and the soil conditions of the area in and around the application site.* This section is to provide reviewers with general background information related to the application site. Include a legal description of the proposal area, the county in which it is located, and as appropriate, distance and direction from the nearest municipality, local landmarks, and named water bodies. Provide elevations and aspect. Describe the underlying parent materials, including their origin (i.e., glacial versus bedrock); the name(s) of any rock formations and their associated characteristics; and geologic structure relevant to slope stability. Describe the soils on site based on existing mapping, field observations, and any available local information. Describe soil texture, depth, and drainage characteristics.

(c) *Describe the potentially unstable landforms of the site.* Include a general description of the topographic conditions of the site. Specifically identify the potentially unstable landforms located in the area (i.e., those defined in WAC 222-16-050 (1)(d)(i)), in addition to any other relevant landforms on or around the site. Describe in detail the gradient, form (shape), and approximate size of each potentially unstable landform. Include a description of the dominant mass wasting processes associated with each identified landform, as well as detailed observations of past slope movement and indicators of instability. Assign a unique alphabetic and/or numeric identifier label to each landform on a detailed site map of a scale sufficient to illustrate site landforms and features. Where the proposal involves operations on or in the groundwater recharge area of a glacial deep-seated landslide(s) specifically discuss the probable impacts to groundwater levels and those impacts to the stability of the deep-seated landslide(s).

(d) *Analyze the possibility that the proposed forest practice will cause or contribute to movement on the potentially unstable slopes.* Explain the proposed forest management activities on and adjacent to the potentially unstable landforms. Clearly illustrate the locations of these activities on the site map, and describe the nature of the activities in the text. Discuss in detail the likelihood that the proposed activities will result in slope movement (separate activities may warrant separate evaluations of movement potential). The scope of analysis should be commensurate with the level of resource and/or public risk. Include a discussion of both direct and indirect effects expected over both the short- and long-term. For proposals involving operations on or in the groundwater recharge area of a glacial deep-seated landslide, conduct an assessment of the effects of past forest practices on slide/slope movement. Explicitly state the basis for conclusions regarding slope movement. Conclusions may be based on professional experience, field observations, unpublished local reports, watershed analyses, published research findings, and/or slope stability model output.

(e) *Assess the likelihood of delivery of sediment and/or debris to any public resources, or to a location and in a way that would threaten public safety, should slope movement occur.* Include an evaluation of the potential for sediment and/or debris delivery to public resources or areas where public safety could be threatened. Discuss the likely magnitude of an event, if it occurred. Separate landforms may warrant separate evaluations of delivery and magnitude. Explicitly state the basis for conclusions regarding delivery. Conclusions may be based on professional experience, field observations, unpublished local reports, watershed analyses, published research findings, and/or landslide runout model results, which should have site specific data.

(f) *Suggest possible mitigation measures to address the identified hazards and risks.* Describe any modifications necessary to mitigate the possibility of slope movement and delivery due to the proposed activities. If no such modifications are necessary, describe the factors inherent to the site or proposed operation that might reduce or eliminate the potential for slope movement or delivery. For example, an intact riparian buffer down slope from a potentially unstable landform may serve to intercept or filter landslide sediment and debris before reaching the stream. Discuss the risks associated with the proposed activities relative to other alternatives, if applicable.

The report should be as detailed as necessary to answer these and any other relevant questions. In particular, examination of aerial photographs (preferably taken over many years) would be appropriate to evaluate the stability characteristics of the area and the effects of roads or previous logging on the subject or similar sites. Field observations will usually be necessary to define the local geology, landforms, etc. Quantitative estimates of site stability produced using SHALSTAB, XSTABL, or other slope-stability models may be useful.

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